

# Glacio-seismotectonics: Ice sheets, Crustal Deformation and Seismicity

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## ABSTRACT

The last decade has witnessed a significant growth in our understanding of the past and continuing effects of ice sheets and glaciers on contemporary crustal deformation and seismicity. This growth has been driven largely by the emergence of postglacial rebound models (PGM) constrained by new field observations that incorporate increasingly realistic rheological, mechanical, and glacial parameters. In this paper, we highlight some of these recent field-based investigations and new PGMs, and examine their implications for understanding crustal deformation and seismicity during glaciation and following deglaciation. The emerging glacial rebound models outlined in the paper support the view that both tectonic stresses and glacial rebound stresses are needed to explain the distribution and style of contemporary earthquake activity in former glaciated shields of eastern

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Canada and Fennoscandia. However, many of these models neglect important parameters, such as topography, lateral variations in lithospheric strength and tectonic strain built up during glaciation. In glaciated mountainous terrains, glacial erosion may directly modulate tectonic deformation by resetting the orogenic topography and thereby providing an additional compensatory uplift mechanism. Such effects are likely to be important both in tectonically active orogens and in the mountainous regions of glaciated shields.

In general, the short-term response to ice fluctuations is similar to the Earth's response to fluctuations in water reservoirs with the subsequent increase or decrease in seismicity which depends on the pre-existing stress state. When the ice fluctuations occur on the spatial scale, and magnitude, of the Late Pleistocene glaciation and deglaciation, however, the viscoelastic response of the Earth (especially the mantle) causes significant changes in crustal deformation and earthquake activity that is spatially extensive and temporally complex. The regions of greatest ice thickness and the regions marginal to the Late Pleistocene ice sheets indicate the most dramatic evidence of earthquake faulting. The mantle response to these large ice mass fluctuations and the change in mass between the oceans and land caused and continues to cause measureable crustal deformation at hundreds of kilometres from the ice margins.

For both tectonically active and cratonic regions, palaeoseismic investigations of Late Pleistocene and Holocene faults are an important tool in evaluating earthquake hazard. The predicted response of the Earth over this time, however, is very dependent on the model assumed. For most regions, far more carefully designed field observations are needed to constrain the existing rheological and ice models.

## 1. Introduction

Since the concept of glacio-isostasy was formulated over a century ago (Jamieson, 1865; 1882; De Geer, 1888), there has been strong theoretical and empirical support for the idea that major glacial advances and retreats are accompanied by significant crustal deformation. Glacio-isostatic observations, particularly of raised marine and glacial lake shorelines, and uplift and subsidence estimates from tide-gauge measurements have been used extensively to develop, test and refine models of Earth rheology (Walcott, 1970; Peltier & Andrews, 1976; Lambeck, 1990). These models demonstrate that glacio-isostatic recovery from Late Pleistocene glaciation is deep seated, involving both elastic flexing of the lithosphere and viscous flow in the mantle, and as a consequence, that it affects large areas. Indeed, crustal deformation related to Late Pleistocene deglaciation is global in extent, due to mantle flow associated with isostatic adjustment, and by eustatic loading of continental shelves and ocean basins due to sea-level rise. Furthermore, even the early workers such as De Geer recognised that the upper-crustal response to ice-mass fluctuations was likely to have involved intense seismic activity (de Geer 1940, cited in Mörner, 1979). The role and significance of seismicity in this process, however, has only been vigorously pursued in the last few decades, driven largely by nuclear-waste disposal and seismic-hazard concerns in the previously glaciated intraplate shields of North America and Fennoscandia (Mörner, 1978; Adams, 1989a; Talbot, 1999). More recently, the recognition that crustal deformation and seismicity are potentially important ‘glacial’ processes has infiltrated into mainstream Quaternary science (Hunt & Malin, 1998).

A modern milestone in the understanding of the interplay between ice sheets, crustal deformation and seismicity was in 1989, with the publication of the proceedings of a NATO workshop on 'Earthquakes at North-Atlantic Passive Margins: Neotectonics and Postglacial Rebound' (Gregersen & Basham, 1989). This collection of papers, albeit with its geographic emphasis on the North Atlantic seaboard, integrated ice-sheet modelling and fault-modelling work, seismological studies, and field investigations of neotectonic phenomena in former glaciated terrains, and in so doing raised important research questions: i) was the final decay of the large continental ice sheets in northern Europe and eastern North America accompanied by a burst of earthquake activity far more intense than those regions witness today? ii) are faults that were activated in postglacial times still active? iii) how do we assess earthquake hazards in deglaciated regions that are now seismically quiescent? iv) what is the relative importance of plate tectonics and postglacial rebound in earthquake generation? This last point is perhaps the most critical.

The last decade has witnessed increasing research into such questions, and in this context, this special issue of *Quaternary Science Reviews* re-examines and updates several of these issues. The volume is developed from a symposium at the XV INQUA Congress, held in Durban, South Africa in August 1999, and demonstrates how the debate about the interplay between tectonic and glacial systems has developed in the last ten years. It addresses issues relating to other passive margins, evaluates the impact of ice-mass fluctuations on crustal deformation at active plate boundaries, and considers the role of glacial erosion on regional-scale landscape evolution, especially uplift. However, it is important to note that the study of the interaction between ice sheets, crustal deformation and seismicity is far broader than the scope of the papers encompassed in this volume. Areas of investigation that are 'missing' from this volume range from studies of the seismicity of glacial dynamics ('icequakes') (Gaul

*et al.*, 1992; Anandakrishnan & Bentley, 1993), and the interaction between glaciation and volcanism. For example, large variations in the eruption rate of the spreading ridge on Iceland over the last 10,000 years led Jull and McKenzie (1996) to examine the possibility that deglaciation caused a higher rate between 8 and 10 kyr BP; we refer the reader to this work for a wider discussion on the effect of deglaciation on mantle melting. Furthermore, there is a growing recognition that glaciation and plate tectonics have combined to shape the Earth over geological time (Eyles, 1993), and that the redistribution of mass that occurs during the growth and decay of ice sheets may perturb the Earth's rotational state. Although the rotational response of the Earth has important implications for climatic variation (e.g. Bills *et al.*, 1999; Peltier, 1998 and references therein) these are not discussed in this collection of papers. Instead, the bulk of the papers in this special issue consider, directly or indirectly, how glacial loading and unloading drives the deeper seated viscoelastic response of the Earth and upper-crustal faulting and earthquake generation. This paper sets out the background to many of the debates currently active within the subject, and attempts to draw out important outstanding issues.

## **2. Definitions and Scope**

Quaternary science has long been concerned with the deformation effects of ice sheets, though mainly from the perspective of providing field or laboratory insights into present-day and, by inference, past glacier regimes. Research into such 'glaciotectonics' conventionally focuses attention on "the structural deformation as a direct result of glacier movement or loading" (INQUA Working group on Glacier Tectonics 1988), and largely considers surficial deformation processes acting beneath or at the margin of glaciers, produced directly by the movement of the glacier (van der Meer, 1987; Hart and Boulton

1991; Harris *et al.*, 1997). By definition these processes are related to the subglacial, and proximal-proglacial domains. By contrast, this volume of papers examines the interplay between glacial dynamics and structural deformation at a crustal scale, and is not concerned with traditional 'glaciotectonics' research as defined by the INQUA Commission (Thorson, this volume). In particular, the bulk of the papers presented here are concerned with the role of postglacial rebound in modulating crustal deformation and seismicity, and some focus specifically on how deglaciation itself is accompanied by marked changes in seismic-strain release patterns - a tendency that has been termed 'deglaciation seismotectonics' (Muir Wood, this volume). In this volume, we use the broader term 'glacio-seismotectonics' to refer to the past and continuing effect of ice sheets and glaciers on contemporary crustal deformation and seismicity.

### **3. Postglacial rebound: new models, old questions**

The repeated growth and decay of major ice sheets cause major changes in vertical load, fluid pressures and crustal strain, thereby potentially influencing the pattern, nature and rate of crustal deformation and seismicity (seismotectonics) in affected regions. The last decade has witnessed a significant growth in our understanding of this process, driven largely by the emergence of new postglacial rebound models that incorporate increasingly realistic rheological, mechanical and glacial parameters (James & Morgan, 1990; Spada *et al.*, 1991; James & Bent, 1994; Wu & Hasegawa, 1996a; 1996b; Wu, 1996; 1997; Johnston *et al.*, 1998; Klemann and Wolf;1998; Wu *et al.*, 1999), as well as a careful evaluation of the model limitations (Davis *et al.* 1999, Wieczerkowski *et al.* 1999; James *et al.* this volume). For example, the recent models of Wu and co-workers incorporate (1) a viscoelastic spherical

Earth model; (2) a realistic ice-sheet history involving repeated cycles of loading and unloading; (3) a more realistic total stress field (comprising glacial/deglacial induced stress, tectonic stress and overburden pressure components); and (4) fault-instability determinations (Wu, 1999). The outstanding contribution of this new generation of rebound models is that they reveal how the viscoelastic response of the mantle to imposed surface loads can exert significant horizontal stresses. As a consequence, glacial rebound is now widely considered as a potential mechanism for crustal deformation and seismicity not only within limits of former ice sheets but also for several hundred kilometres beyond the limits of glaciation (James and Bent, 1994). Although these models are discussed in more detail by Muir Wood (this volume), the key elements of how glacial loading/unloading cycles influence crustal seismotectonics are summarised here.

Prior to glaciation, crustal deformation proceeds at a rate and with a style that is driven largely by horizontal plate motions, but with the onset of glaciation the prevailing tectonic stress state is overprinted by a glacial stress stress (Fig. 1). Fluctuating glacial loads result in a lagged crustal response because while the elastic response to unloading/loading is instantaneous, the viscoelastic response of the mantle is slower. Furthermore, the density contrast between ice and rock ensures that the magnitude of the crustal adjustment is less than the imposed ice-mass fluctuations (Fig. 2). Elastic crustal flexure results in a 'bowl' of depression below the ice-sheet centre, accommodated by crustal contraction as deeper mantle material flows radially outward from below the maximum ice load (Fig. 3a). Horizontal crustal motions induced by the ongoing glacial isostatic adjustment increase out from the ice-sheet centre and are greatest at the ice-sheet margins (James and Bent 1994; Peltier 1997). In the upper crust, flexural stresses imposed by an ice load of a few km thick will be a few tens of MPa (Walcott, 1970; Stein *et al.* 1989), which may typically match or exceed the

magnitude of regional tectonic stresses (e.g. the present-day maximum horizontal compression [ $\sigma_H$ ] in eastern North America is  $\sim 10$  MPa (Adams 1989b)) (Fig. 1).

Beyond the ice-sheet margins, in the ice-free foreland, upper-crustal flexural upwarping by the adjacent glacial load, combined with the deep-seated in-migration of sub-crustal material exuded from below the ice sheet, create uplift and radial crustal extension within the surrounding forebulge rim (Fig. 1). Extensional stresses of up to 10 MPa are predicted for a forebulge related to a roughly 1 km thick ice load (Walcott 1970, Stein *et al.* 1989). During the same glaciation phase, the global lowering of sea level may also contribute to eustatic unloading of continental shelves.

On deglaciation, the abrupt removal of the vertical stresses imposed by ice loading initially leads to elastic rebound, causing crustal uplift, faulting and seismicity; the horizontal stresses are more gradually relaxed through deeper viscoelastic return flow of the mantle material (Fig. 3b). The result is a rising dome of radial crustal extension in the centre of the former ice sheet that expands outward, to 'recover' a forebulge that is collapsing under crustal contraction. In seismotectonic terms, the deglaciated upper-crustal region should be in tension (normal faulting) and the unglaciated forebulge should be in compression (thrust faulting), though compressive stress regimes often prevail within formerly glaciated areas due to ongoing regional tectonic compression. Nevertheless, across the margins of former ice sheets there are often marked changes in faulting style. For example, the focal mechanisms of the larger events indicate that thrust faulting is the dominant mechanism in Eastern Canada. However, along the northeast coast of Baffin Island, focal mechanisms are of the normal-



fault type while in northeastern United States, strike-slip faulting predominates. (Stein *et al.*, 1989, Hasegawa and Basham, 1989).

Models suggest that for a 1 km thick ice load, the rebound is several hundred metres, and the corresponding stresses caused by deglaciation flexure typically reach tens of MPa, which is equivalent to plate-driving stresses but may be almost an order of magnitude less than stresses imposed by sediment loading on adjacent continental margins (Stein *et al.*, 1989). Within the former ice-sheet centre and its forebulge, predicted horizontal strain rates imposed by glacial rebound may be several orders of magnitude greater than observed seismic strain rates (James and Bent, 1994). At greater distances from the load centre, crustal subsidence occurs due to loading of the continental shelves and ocean basins by water returned to the global oceans (Walcott, 1970; Peltier, 1976; Lambeck, 1993).

The solid Earth relaxes towards a state of isostatic equilibrium long after the ice has melted. However, because the northern hemisphere has been subjected to repeated continental-scale glaciations, each of around  $10^4$  yr duration, during the Pleistocene (Shackelton *et al.*, 1984), it is probable that the cumulative effects of multiple glacial loading/unloading cycles have contributed to the present-day glacially-induced strain field in this region (Klemann & Wolf, 1998).

#### **4. How realistic are ice-history/Earth models?**

The specific response of the Earth to a change in surface load, across different spatial scales and on a variable time scale, depends on the rheological structure of the Earth. The effective

viscosity structure within the mantle and thickness of the overlying rigid layer determine, to a significant extent, the timing, pattern and magnitude of postglacial surface displacements.

Over the past 30 years,  $^{14}\text{C}$  dated palaeoshorelines, and the recent sea-level record, have been used to infer the rheological properties of the Earth and the space-time geometry of the Late Pleistocene ice cover (Walcott, 1970; Peltier, 1976; Lambeck 1993). Specific ice history/Earth models derived from such data, however, are both non-unique and the global models do not account for lateral variations in rheology (see discussion by Davis *et al.*, 1999).

As a consequence, there remain a number of important questions that need to be addressed: i) how does the crust and mantle rheology control the spatial and temporal characteristics of surface deformation following glacial fluctuations? ii) how much variation is there between rheological structure in plate-boundary regions and plate-interior regions? iii) what is the 'trade-off' between lithospheric thickness and asthenospheric viscosity for determining the predicted surface deformation? iv) what is the 'trade-off' between ice-sheet models and Earth rheology in predicting surface displacements? v) what kinds of additional high-resolution field data are needed to constrain rheological models?

Global rheological models do not include lateral asthenospheric viscosity variations, though these would be especially important in calculating deformation rates due to postglacial rebound in tectonically active regions (Kaufman and Wu, 1998). James *et al.* (this volume) used shoreline tilts of two proglacial lakes and rapid sea-level fall following the retreat of the Cordilleran ice sheet at the northern Cascadia subduction zone to constrain the response of the Earth to ice unloading. Their paper shows the trade-off between the effective viscosity of

the upper mantle and the thickness of the lithosphere. To account for the uplift observations, an upper mantle viscosity of less than  $10^{20}$  Pa s is required. This is lower than viscosity estimates derived from postglacial rebound studies of tectonically less active regions. There is a growing body of results from other tectonically active regions which suggest significant regional variability in rheological structure (Iceland, Sigmundsson and Einarsson, 1992; Japan, Okuno and Nakada, 1998; Basin and Range of the United States, Nakiboglu and Lambeck, 1983; Bills and May, 1987). The lower viscosity values for the upper mantle near the Cascadia have implications for both the temporal and spatial pattern of postglacial rebound, as well as implications for modulating the timing of large earthquakes.

Global postglacial rebound models such as ICE-4G (Peltier, 1994) make predictions of global sea-level for the last 11,500 calendar years that can be compared with Holocene sea-level change histories for particular coastlines (Shennan *et al.*, 2000). To investigate sea-level changes over a longer time span, Rostami *et al.* (this volume) used Quaternary marine terraces along a 1000 km long stretch of the east coast of Argentinian Patagonia. The inferred Holocene sea-level histories were compared to those predicted from the ICE-4G model of the global process of glacial isostatic adjustment following Late Pleistocene deglaciation (Peltier, 1994). Although the ICE-4G model predictions match well with the documented relative sea-level (rsl) history along the northern part of the east coast of South America, Rostami and co-workers found a misfit between ICE-4G predictions and observed rsl history in Argentinian Patagonia. They hypothesized that the wide Argentinian shelf could cause a hydro-isostatic response to sea-level fluctuations that is not accounted for in ICE-4G, or that significant Holocene uplift of the region associated with subduction beneath South America systematically distorts the marine terrace heights. Another important effect in regions with significant recent ice cover, such as Patagonia and Alaska, is the influence of

late Holocene glacial fluctuations on uplift rates and seismicity (Ivans and James, 1999, Sauber *et al.*, 2000).

## **5. Do large ice sheets suppress seismic activity?**

An apparent tendency for large continental ice sheets to suppress tectonic activity was first noted by Johnston (1987, 1989). According to Johnston, in regions where the maximum tectonic stress is horizontal (that is, in strike-slip and compressive tectonic regimes, but not in tensional regimes), the existence of an ice sheet will tend to increase vertical stresses far more than horizontal stresses, thereby bringing vertical and horizontal stresses more into balance (Fig. 3a). The resulting reduced ‘deviatoric stress’ promotes the conditions for fault stability, and therefore, inhibits seismic-strain release. However, Muir Wood (1989) recognised that ongoing tectonic motions will tend to progressively reduce crustal stability the longer the ice sheet is in existence. Where tectonic loading was fast enough in relation to the duration of the ice sheet (as in plate boundary regions), then the crust below the ice sheet would eventually revert to its pre-glacial tectonic state, and seismicity would resume at levels comparable to those that existed before ice-sheet growth (Muir Wood, this volume). Johnston (1987, 1989) argued that glacial loading could explain the relative aseismicity of the crust beneath the present-day ice caps of Antarctica and Greenland. It is possible, however, that these areas have, in fact, adjusted to their glacial loads and that they are simply intraplate areas with less seismic-strain release. In short, the suppression of seismicity by ice sheets need not apply to every glaciated or formerly glaciated region, as is demonstrated in the following discussion which shows that the behaviour of large ice sheets need not apply to smaller ice sheets, whose imposition can actually promote crustal instability (Johnston *et al.*, 1998).

## 6. Do great ice sheets induce large earthquakes?

Overall, the combination of deep-seated faults cutting cratonic shields formerly loaded by major ice sheets may be the ingredients for earthquake magnitudes that are far greater than can be hosted by other intraplate regions. In layman's terms, it seems as if great ice sheets can induce large earthquakes (Johnston, 1996). However, recent modelling studies suggest that the size of ice sheets may not be the main control on the degree of crustal instability they promote.

At the culmination of the Last Glacial Maximum, ice sheets had thicknesses in the range of 2-3 km (Boulton, 1985) and radii approaching 1000 km in Fennoscandia and 2000 km in Laurentia. Ice sheets in Iceland, the British Isles and the mountain chains of Europe were smaller and thinner. Recent modelling suggests that these different scales of glacial load may have induced very different crustal responses. In particular, Johnston *et al.* (1998) have shown that the patterns and magnitudes of stress in a flexed lithosphere depend on the dominant wavelength of the load (twice the ice-sheet diameter) relative to the elastic thickness of the lithosphere. Greatest stress changes are induced by ice loads with dominant wavelengths close to 12 times the elastic thickness of the plate, when horizontal stresses may be up to six times as large as vertical stresses. According to the Earth-model parameters used by Johnston *et al.* (1998), horizontal stresses are amplified to their greatest extent for ice loads with a radius of ca 280 km, which is roughly comparable to the dimensions of the main British ice sheet at the Last Glacial Maximum. For a modelled ice sheet of 333 km in radius and 1000 m thickness, horizontal stresses are predicted to be up to 2.5 greater than the

vertical stress (23 MPa and 9 MPa, respectively). For larger ice sheets (with radii of ca 1000 to 2000 km), the amount of horizontal stress is of a similar magnitude to the vertical stress, since the dominant wavelength of the larger ice sheets is further from the critical wavelength (though since more extensive ice sheets are invariably also thicker, horizontal stress changes may be much larger). The modelling supports the contention by Johnston (1987) that for glacial loads of large lateral extent (relative to lithosphere thickness), ice-sheet loading stabilises faults underneath the load, while unloading destabilises faults there. For small (radius ca 333 km) ice sheets, however, loading causes an increase in stability of less than 1 MPa at shallow depth, but promotes instability at greater depths. Beyond the ice margin, loading decreases stability for all ice sheets, and unloading also decreases stability relative to the initial pre-glacial state.

Several important implications arise from such models. Firstly, whereas stability is expected beneath larger ice sheets, fault instability is promoted underneath moderate to small ice sheets. Secondly, the largest ice sheets need not generate the largest earthquakes, in part because this depends on the pre-existing stress state. In addition, Johnston *et al.* (1998) argue that because the lateral extent of the 1000 km radius Fennoscandian ice sheet was closer to the thickness of the lithosphere than the 2000 km radius Laurentide ice sheet, larger horizontal stresses, and therefore greater potential rebound stresses to reactivate faults and trigger earthquakes, existed for the smaller Fennoscandian ice load. Thirdly, because ice sheets grow and decay in extent during a glacial cycle, we should expect their consequent effect on crustal stresses to similarly change over time. For example, Hasegawa and Basham (1989) suggested that a receding Laurentide ice front at 10-9 kyr BP could have reactivated normal faults along the St Lawrence seaway, while, at 8 kyr BP, the ice margin was positioned appropriately to trigger faulting in the Peel sound area of Arctic Canada (Fig. 4a).

More recently, Talbot (1999) has proposed that the pulse of seismotectonic activity in the Lapland Fault Province may have coincided with the time when the Late Weichselian ice sheet in that region was of a size that coincided with the critical wavelength of the underlying lithosphere. The same might be expected to have occurred during the decay of the Laurentide ice sheet. Fourthly, the models highlight how glacial loads are unlikely to act in isolation. For example, the influence of the Fennoscandian ice mass is predicted to have led to greater crustal stability for the British Isles relative to other areas in its peripheral forebulge, even though the main British ice sheet acting alone ought to promote crustal instability below it (Johnston *et al.*, 1998). Again, we might also expect there to have been similar destructive (or alternatively, constructive) interference of opposing glacial stress fields in eastern North America, where different major ice lobes co-existed during the Last Glaciation (Fig. 4a). Around 8 kyr BP, for example, the Boothia Uplift-Bell Arch area, a prominent seismotectonic zone, was located in the cumulative forebulge of three separate loading centres, which we would expect to reinforce their crustal effects (Hasegawa and Basham, 1989).

The focus of most post-glacial rebound modeling is on the retreat of the extensive ice sheets, but the question remains as to whether fluctuations in loading by smaller ice masses have a discernible seismotectonic influence? And if so, over what time and spatial scale are discernible effects observed. Johnson *et al.* (1998) modeled the crustal effects of ice sheets as small as 330 km in radius. They found that for smaller loads there is faster relaxation because the smaller load is partly supported by the elastic lithosphere and it mostly stresses the less viscous upper mantle, which flows faster than the more viscous lower mantle. However, empirical studies suggest that the smaller ice sheets may have a continued influence on seismicity. In Britain, Musson (1996) noted that much of the present-day seismicity coincides with the former margins of the Younger Dryas ice cap which had a

radius of only ca 50 km covering part of the western Scottish Highlands for a period of about 1000 years, following the decay of the LGM British ice sheet. Musson speculated that optimally oriented faults that were reactivated during decay of the LGM ice sheet may remain seismically active. Rebound caused by the deglaciation of a moderately small ice cap that covered the Swiss Alps at the Last Glacial Maximum may still be a large component of the present-day uplift rate detected in that area (Gudmundsson, 1994). Significant ongoing glacial rebound in the Swiss Alps may explain enhanced early Holocene seismic activity (Beck *et al.*, 1996), postglacial faulting (Geiger *et al.*, 1986) and even large historical earthquakes (Meyer *et al.*, 1994) in the region.

Theoretical models and field studies in southern Alaska show that changes in the thickness of ice due to recent glacier surges, as well as glacial retreat this century, can produce measureable crustal deformation on a local and regional scale (Clark, 1977; Cohen, 1993; Sauber *et al.*, 1995, 2000). Additionally, Sauber *et al.* (2000) show that the stress changes associated with these glacial changes are significant relative to the stress drop in earthquakes and may cause fluctuations in the background seismicity. The suggestion by Cohen (1993) that recent ice mass redistribution in regions which are not tectonically active might also produce locally significant and detectable crustal deformation, and thereby a seismotectonic expression, remains untested. For the glacier surge case, the spatial extent of the ice-mass fluctuations may be less than the thickness of the lithosphere and the Earth would respond primarily as an elastic medium. The seismotectonic effect would be similar to observed changes associated with water-reservoir impoundment with changes in local and even regional seismicity (see discussion by Thorson, this volume). The global retreat of temperate glaciers this century occurs on a more regional scale in places like Alaska and Patagonia, and the viscoelastic response of the Earth needs to be considered. Recent ice-



mass fluctuations of the Greenland ice sheet as well as Antarctica are the subject of ongoing studies (Wahr *et al.* 1995; James and Ivins, 1998, Van Dam *et al.*, 2000).

## **7. Was deglaciation a period of enhanced seismotectonic activity?**

A decade ago, Gregersen & Basham (1989, p.3) concluded that “...there has been a growing consensus among seismotectonic specialists that in glaciated regions the rates of significant earthquake activity were far greater immediately following deglaciation than they are today.” This view largely stemmed from the emergence in the 1980s of evidence for significant crustal faulting and palaeoseismicity in northern Fennoscandia, and to a lesser extent eastern Canada, around the time of deglaciation (Lagerbäck and Witschard, 1983; Muir Wood, 1989). Thus, in the Lapland Fault Province of northern Fennoscandia (Fig. 5), geomorphological and stratigraphical studies indicated that all the main postglacial faults are likely to have formed in association with deglaciation (Muir Wood, 1989; 1993; Talbot, 1999). For example, relations between the 50 km long Lansjärv Fault and different landforms produced during ice wastage (e.g. eskers, meltwater channels and shorelines) and a complex till stratigraphy revealed by fault-trenching, provided a good reference for dating fault movement, and showed that faulting occurred within a few years to decades of being uncovered by the ice (Lagerbäck, 1988; 1990; 1992). Similarly, the 165 km long Pårve Fault was shown to have developed while ice still covered part of the fault’s trace (Lagerbäck & Witschard, 1983). These and related studies indicate that, in Fennoscandia, faulting can be directly linked to the final stages of ice-sheet recession, with the bulk of this seismotectonic activity probably occurring within a very short period, possibly only a few hundred years

(Muir Wood, 1989; 1993). In this context, Talbot (1999) referred to the Lapland faults as 'endglacial', and argued that they may be a product of the rapid deglaciation of this region.

Recent modelling work has largely corroborated this view, though the models reveal how the locus, timing and duration of this 'pulse' of deglaciation seismotectonics may vary dramatically both between different glaciated regions and across a particular region. For example, according to Wu (1999), in the central part of the Laurentian ice-sheet, fault instability begins at around 7-9 kyr BP and the greatest instability is reached around 7-4 kyr BP. In ice-marginal areas, such as in Newfoundland, where ice wastage occurred earlier, instability is initiated earlier, around 13 kyr BP, but was likely to have been suppressed during the period 10-7 kyr BP when rapid ice wastage caused strong eustatic loading of continental marginal locations. In areas beyond the ice margin, such as in the northern U.S.A., the onset of seismic and fault activity was delayed until around 8 kyr BP. For Fennoscandia, the models predict the onset of postglacial fault activity is around 11-9 kyr BP, prior to complete deglaciation, and maximum instability is attained around 10-7 kyr BP, far earlier than in North America (Wu, 1999).

These models reveal how at any given time, different parts of a glaciated domain will behave differently. For example, at around 9 kyr BP, during the early stages of deglaciation in Laurentia, crustal stability is promoted both underneath the existing ice and further away in the ice-free forebulge, whilst instability is concentrated in the intervening deglaciated ice-marginal zones. Such model predictions, however, are highly sensitive to the viscosity structure used, and ideally could be validated and refined against high-resolution field data on

the timing and duration of fault activity in different regions. Unfortunately, as is discussed later, no such dataset currently exists.

### **8. Are the Lapland postglacial faults a 'unique' tectonic environment?**

In the Lapland Fault Province, several of the postglacial faults have lengths of 50-150 km and maximum surface displacements of 5-30 m; in several cases, displacements of up to 15 m are considered to have been achieved in one, or at most two, slip events (Muir Wood 1989) (Fig. 5, Table 1). Postglacial faults like the Pårve Fault appear to cut the entire thickness of the Fennoscandian crust (Arvidsson, 1996), which Slunga (1989) showed is seismogenic down to the Moho near 45 km. Based on their inferred fault-displacement / fault-area dimensions, the Lapland faults may have hosted earthquakes with magnitudes as high as 8.2 (Pårve Fault) and 7.8 (Lansjärv Fault) (Johnston 1996) (Table 1). Much of our understanding of glacio-seismotectonics comes from northern Fennoscandia, but are the seismotectonics of this region typical of deglaciated regions more generally?

It remains uncertain whether faults comparable to those in the Lapland Fault Province are characteristic of other formerly glaciated terrains, or whether this instead represents a 'unique tectonic environment' (Muir Wood 1989). Such large faults may simply not exist in areas like eastern North America, where rebound stresses associated with the Laurentide ice sheet are predicted to be less than the Fennoscandian ice sheet (Johnson *et al.*, 1998). Adams (1989) and Muir Wood (1989) had speculated that, if they were to occur at all, broadly equivalent glacial rebound conditions might be expected along the northeastern margin of the former Laurentian ice sheet, between coastal Labrador and Baffin Island. Ten years on, there remains

no evidence for significant postglacial faults in these areas. An inventory of postglacial faults in eastern and Arctic Canada (Fenton 1994) reveals only two candidate structures that might possibly approach the scale of the Lapland faults; (1) the Aspy Fault in eastern Canada, and (2) the Peel Sound fault zone in Arctic Canada. The Aspy Fault offsets a 125,000 year old interglacial rock platform on Cape Breton Island (A in Fig. 4a), but the fault scarp is not exposed and surficial deposits show no evidence of Holocene movement (Grant, 1990). In the Peel Sound area of Northwest Territories (P in Fig. 4a), elevated 9300 year old shorelines appear to be offset across the sound by an intervening fault or group of faults with a cumulative throw of 60-120 m (Dyke *et al.*, 1991). The absence of dislocations on younger shorelines brackets the duration of fault activity to an interval of around 1000 years during deglaciation, after which no significant fault movement occurred (Dyke *et al.*, 1991), but the causative fault zone appears to lie offshore, away from direct investigation. Conceivably, both these faults may have hosted large single-event surface displacements comparable to the Lapland faults, but there are no signs of the Aspy Fault having been activated during the most recent deglaciation and shoreline dislocations in the Peel Sound can not be related to specific faults.

Based on our current understanding, therefore, postglacial faults in the Lapland Fault Province appear to have much larger displacements, and much greater lengths, than the Canadian equivalents. In both regions, however, most of the reported postglacial faults are compressional structures (thrust faults), a mode of failure that is consistent with a glacial rebound mechanism (Wu, 1999). However, glacial rebound models also predict much spatial and temporal variability in the style of seismotectonic activity during and following deglaciation (Fig. 6). At the Last Glacial Maximum (20 kyr BP), for example, normal faulting is predicted below the Fennoscandian ice cover, but pockets of thrust faulting and strike-slip

faulting may have co-existed with this. Following deglaciation (8 kyr BP), thrust faulting dominated within the former glaciated region and normal faulting dominated the forebulge, but these are considered to have been separated by belt of strike-slip faulting that becomes more extensive into late Holocene times. Large strike-slip faults are suspected to have been active in Fennoscandia (Talbot and Slunga 1989), but they have not been demonstrated to have appreciable postglacial offset. However, significant ( $10^1$ - $10^2$  m) horizontal displacements are reported for short (1 - 14 km long) faults in the Scottish Highlands (Ringrose, 1989a; Fenton, 1992), which lay both in the forebulge of the Fennoscandia ice sheet and below the much smaller main British ice sheet. In this volume, Firth & Stewart question the evidence for significant postglacial strike-slip faulting in Scotland, and Fig. 6 suggests that thrust faulting was maintained in this region until late Holocene times. Nevertheless, the message of Fig. 6, and also of Fig. 1, is that we can expect a diversity of structural styles and changing modes of fault failures in different parts of a deglaciated region, and indeed at the same location at different points in time. Clearly, the expression of glacio-seismotectonic activity in areas away from regions of major ice-mass fluctuations is going to be far more varied and complex than the dramatic evidence of earthquake faulting preserved in the Lapland Fault Province. How then, do we recognise the surficial effects of glacio-seismotectonics?

## **9. How can we recognise postglacial seismotectonics?**

Identifying field evidence for postglacial seismotectonic activity is problematic, since the surficial effects of ground rupture and seismic shaking in former glaciated terrains may be difficult to distinguish from glaciogenic, cryogenic or loading phenomena. In both eastern

North America and southern Fennoscandia, faults affecting glacially-scoured surfaces have been interpreted by some workers as the neotectonic structures (Adams, 1989b; Mörner 1990) and by others as the products of glacial erosion and transport (Broster & Burke, 1990; Talbot, 1999). In some cases, the difficulty in distinguishing deformation structures caused by glacier over-riding from neotectonic structures have important seismic-risk implications (Mohajer *et al.*, 1992; Adams *et al.*, 1993). So how we can discriminate between tectonic deformation and non-tectonic deformation processes?

Currently, postglacial faulting is identified on the basis of key 'diagnostic criteria' that are argued to discriminate neotectonic surface faulting from other unrelated phenomena (Mohr, 1986; Fenton, 1992; 1994; 1999; Muir Wood, 1993). In assessing the claims of neotectonic faulting in Norway, for example, Dehls *et al.* (this volume) use the scheme of Muir Wood (1993), in which key criteria are: (1) offset of a surface or material that was originally continuous and unbroken, and whose dislocated fragments, if dated, can be demonstrated to be of the equivalent age; (2) offset can be shown to relate directly to a fault, and not the result of differential compaction or drape over pre-existing (erosional) scarps; (3) the ratio of displacement to overall length of the feature is more than 1:1000; (4) fault displacement is reasonably consistent along the length of the feature; and (5) fault movement is synchronous along its length. On the basis of satisfying these criteria, suspected postglacial faults are awarded varying degrees of uncertainty: (A) Almost certainly neotectonics; (B) Probably neotectonics; (C) Possibly neotectonics; (D) Probably not neotectonics; and (E) Very unlikely to be neotectonics. Although this scheme is widely applied to assess claims of neotectonic faulting in Fennoscandia, it is debateable as to the extent to which such simple checklists can effectively consider the myriad of different scales and styles of postglacial deformation that

may occur in all former glaciated regions. Essentially, however, there are two important attributes in recognising postglacial tectonic faults - their age and their tectonic mechanism.

As yet, no postglacial fault in eastern North America or Fennoscandia has been dated radiometrically to give an absolute age for a paleoearthquake/faulting event. Instead, the ages of postglacial faults are deduced from their relations to glaciogenic deposits and landforms that, in turn, track the local deglaciation history. In this respect, a postglacial fault ought to demonstrably disturb and/or displace Lateglacial or Holocene sediments or morphological features, such as shorelines or glaciated surfaces, and ideally ought to exhibit no evidence of glacial modification (Fenton, 1992). In this regard, both Mohr (1986) and Fenton (1992) contend that trenching is generally necessary to demonstrate that a scarp is unequivocally the result of Lateglacial or postglacial fault activity and qualify the degree of glacial modification.

In terms of establishing a tectonic origin, we might expect that a postglacial fault ought to be comparable in style to recent earthquake faults rupturing other stable continental interiors. The surface expression of the fault is likely to be a misleading indicator, since postglacial faults may vary markedly in their displacement, geometry and morphology along strike (Muir Wood, 1993), and we know from modern earthquake ruptures that consistent slip along tectonic faults rarely occurs (stress fields are so heterogeneous on the local scale). An alternative approach is to consider that the parameters of postglacial faults (fault displacement, fault length) ought to be comparable with those documented from earthquake faulting in other stable continental regions (Wells and Coppersmith 1994). In particular, fault displacement : length ratios of 1:10,000- 1:100,000, which is typical of reverse-fault

intraplate earthquakes ruptures, are generally considered as indicating neotectonic faults (Muir Wood, 1993). In the Lapland Fault Province, structures like the Pärve and Stuoraggura Faults have displacement : length ratios in this range, but other faults (such as the Lansjärv Fault) have considerably higher (1:1000) ratios; it is still unclear what these higher ratios mean (Muir Wood 1993). They may be signifying different causative mechanisms. High displacement-to-length ratios, for example, are reported for postglacial faults in the French Alps, the Cantabrian Mountains (NW Spain) and the northern Andes that have been attributed to gravitational spreading (i.e. sacking) or localised postglacial rebound (Alonzo and Corte, 1992; Ego *et al.*, 1996; Sébrier *et al.*, 1997). However, Kanamori and Allen (1996) present empirical evidence that suggest that earthquakes with long repeat times generally have higher stress drops and, thereby, might be expected to generate faults with high displacement : length ratios. Thus, high displacement : length ratios may simply indicate that faults may not have been 'activated' since the last glacial event or may only be very slowly accumulating tectonic strain. Since displacement : length ratios have been used to estimate palaeo-magnitudes (Fenton, 1992), a much larger dataset of faulting events in former glaciated regions is required to establish if such fault parameters are sensitive and reliable indicators of tectonic faulting in former glaciated regions (Dehls *et al.*, this volume). A crucial uncertainty is how slip on glacially-activated faults varies as a function of depth. For example, does heightened shallow crustal stresses cause displacement to be greatest at the surface rather than at the base of the seismogenic layer in tectonic faults?

Dehls *et al.* (this volume) present a high-quality case study of two long-suspected postglacial faults (Stuoragurra and Nordmannvikdalen) in northern Norway. The study demonstrates how a combination of geological and geomorphological mapping, geophysical imaging, and in one case, trenching across the possible fault-line, is needed to demonstrate postglacial



displacement and rule out other non-tectonic mechanisms. The results of this integrated approach strongly support postglacial faulting, but the draw attention to many of the problems involved in this type of research. Firstly, the structure revealed by the trench has many characteristics similar to the internal structure of a turf-banked lobe developed by gelifluction over permafrost in a periglacial region (French, 1996). In particular, Figures. 3 and 4 which record the structures in the cross-sections revealed by trenching, show forms that are very similar to the overturned sediments of a surf-banked lobe, caused by caterpillar-like downslope movement of the surface material (Harris, 1972; Elliott and Worsley, 1999). The seismotectonic interpretation proposed by Dehls *et al.* is confirmed only by their high-quality, detailed representation (see their figure 5) of the sedimentary structures exposed by trenching, which show abundant load structures (what they call convolutions) of a type that would not have been formed by gelifluction. Similarly, the landform identified by Dehls *et al.* has a form that is characteristic of mass-movement features, especially turf-banked lobes, formed in periglacial regions. Indeed, close scrutiny of Figures 9 and 11 shows that small lobate forms extend downslope from the general line of the fault displacement, suggesting that some mass movement has occurred since faulting. This highlights a real problem of studying landforms in regions that have been glaciated. Namely, the development of permafrost and periglacial mass-movement following deglaciation which results in the formation of landforms that are similar to those formed by tectonism. This problem is addressed by Dehls *et al.* and they emphasize the linear nature of their faultline, which would not form by periglacial mass-movement, and the small scale of the lobate displacements, which reflect the effects of mass-movement since the tectonic event. Finally, investigations of landforms in periglacial regions emphasize the problems of dating buried soils. Matthews (1980) revealed that ages from a single buried soil (a podzol) ranged over 2250  $^{14}\text{C}$  yrs due

to mixing and the residence time of organic materials in soils - a problem clearly relevant to the dating of postglacial faulting.

The study of the Stuoragurra Fault by Dehls *et al.* suggests that the fault had around 10 m of displacement, and that this was achieved in one major faulting event. Preliminary investigations of the Nordmannvikdalen fault are consistent with this being a tectonic structure, but a gravitational origin can not be eliminated for the reasons outlined above. In both cases, however, the timing of fault activity is only loosely constrained, though both probably formed shortly after deglaciation. Alongside the Lansjärv and Pårve faults, the Stuoragurra Fault is now one of the most intensively studied postglacial faults, but key aspects of its movement history still remain uncertain. The study both exemplifies the multidisciplinary investigations that will be increasingly expected to be applied to the study of postglacial faults, and the likelihood that, even with these, much of the evidence for past surface faulting in former glaciated terrains remains ambiguous and suggests the need for new innovative approaches.

In lamenting the poor surface expression of the 1989 Ungava surface rupture, Adams *et al.* (1991) suggested that, despite the indirect nature of the evidence, a disturbed horizon in adjacent lake sediments may provide the best long-term record of that earthquake, and perhaps of other earthquakes on the Canadian shield. Sediment liquefaction and ground failure phenomena described from recent moderate (M 5-6) earthquakes in formerly glaciated settings (Tuttle *et al.* 1990; Davenport *et al.* 1994) remind us that even comparatively modest earthquakes may trigger widespread and permanent ground effects. Such case studies provide (1) key diagnostic criteria for recognising the seismic origin of relic liquefaction or landslide phenomena, and (2) the basis for empirical relations that permit estimation of past

earthquake magnitudes and shaking intensities (see recent reviews by Jibson, 1996, and Obermeier, 1996). However, care must always be taken to ensure that these deformation structures are not produced by simple loading of sediments deposited rapidly, with an inverse density gradient, in conditions that maintained very high pore-water pressures (see Allen, 1984, Volume II, Chapter 9 for a comprehensive review of the processes). Often it is impossible to differentiate between a simple gravitationally-induced structure and a shock-induced structure, but examination of the sedimentary sequence with respect to the density gradient can be helpful. Where vertical sequences deposited in water, have more-dense material (sands, gravels) overlying less-dense material (silts, silty clays that hold large volumes of water) there would be a tendency for gravitationally-induced deformation (loading) to take place, whereas sedimentary sequences with a normal density gradient are likely to require some independent cause, such as a shock provided by an earthquake, or a glacier ploughing to a sediment body.

With caution, derived from an understanding of the conditions that determine sediment deformation, it is possible to use marine and lacustrine sediment sequences to serve as an archive of palaeoearthquake activity. This requires the preservation and identification of unequivocal coseismic 'event horizons'. The nature of this evidence can take several forms. Anomalous or distinctive lake sediment horizons have been correlated with historical earthquakes in eastern Canada (Doig, 1990, 1991, 1998) and the European Alps (Chapron *et al.*, 1999). Submarine geophysical and coring investigations have identified probable earthquake slumps and flows in lake sediments in Canada (Shilts & Clague, 1992; Ouellet, 1997) and the French Alps (Beck *et al.*, 1996). Deformed Lateglacial or postglacial laminated sediments have been interpreted as earthquake-induced structures in Scotland (Davenport and Ringrose, 1987; Ringrose, 1989b), Sweden (Lagerbäck, 1991; Mörner, this volume) and

eastern Canada (Broster & MacDougall, 1997). In such environments, the ideal conditions for identifying palaeoseismic events are laminated sedimentary sequences, where specific laminae can be related to a varve chronology against which it is possible to determine the age of the disturbance. Recent studies utilising this approach in Fennoscandia include those by Mörner & Trötfen (1993), Mörner (1996, this volume), and Trötfen & Mörner (1997), and a general overview of the method is given in Trötfen (1997). Offshore, high-resolution marine depositional sequences also have the potential to record earthquake-triggered slumping from adjacent continental margins (Grantz *et al.*, 1996).

In most of the above cases, multidisciplinary investigative practices (high-resolution seismic, side-scan sonar and borehole drilling) and detailed analyses of the sedimentological, geochemical and palaeoenvironmental characteristics of the disturbed sequences are needed to establish the likelihood of the seismic hypothesis. The rapid emergence of such studies over the last decade suggests that this will be an important areas of future palaeoseismological research, and one in which that Quaternary scientists can make important contributions.

#### **10. Are faults that were activated in postglacial times still active?**

“Available evidence suggests that the forces that produced large fault offsets in the immediate postglacial period do not present a particular hazard today. The Fennoscandian reverse faults, for example, are essentially aseismic.” (Gregersen & Basham, 1989, p. 3). But is this really the case? Faults are lines of weakness that may be persistently reactivated, so their likelihood of reactivation is largely dependent on whether the prevailing state of stress has brought them close to failure, and whether their orientation, with respect to the prevailing

stress regime, allows them to move. In short, faults that are favourably oriented with respect to the contemporary stress regime may be 'potentially active' (Muir Wood & Mallard, 1992). In eastern North America, a marked rotation in the direction of greatest horizontal compressive stress since early postglacial times (Fig. 4a) (Adams, 1989b; Wu, 1997) suggests that faults reactivated during deglaciation are no longer likely to be active. However, the 1989 Ungava earthquake in northern Canada reactivated a NE-trending thrust (Adams *et al.*, 1991), consistent with the style and trend of the postglacial faults, but apparently at odds with the prevailing contemporary stress regime (Fig. 4a). In northern Sweden, a recent microseismic study has shown that the Pårve Fault remains seismically active (Arvidsson, 1996), and there appears to be some general spatial correlation between seismicity and mapped postglacial faults (Bungum & Lindholm, 1997), including the Stuoragurra Fault discussed by Dehls *et al.* (this volume). Despite this, each of the main Lapland postglacial faults are interpreted as the result of a single 'endglacial' slip event, and there is no surficial evidence for their steep scarps having been destabilised by later seismic shaking events (Talbot, 1986, pers. comm.). In short, it remains uncertain as to how 'weak', and thereby 'potentially active', the postglacial faults in Lapland are. Nevertheless, glaciated cratonic areas like Fennoscandia and eastern Canada are riddled by pre-existing faults and shear zones, many of which appear optimally oriented for reactivation in the present-day stress regime, so recognising possible postglacial faults is itself a relatively ineffective way of identifying potential seismogenic zones.

In many tectonically active regions, palaeoseismic investigations of Late Pleistocene and Holocene faults are becoming an important way of shedding light on their likely future behaviour. In less active glaciated regions, however, such palaeoseismological investigations face particular challenges. The tendency of ice sheets to inhibit seismic activity and to

remove signs of seismotectonics formed before the glacial event means that, almost inevitably, the time window available for palaeoseismology extends from the onset of deglaciation to the present day, a period roughly corresponding to the last 16-7 ka depending on the location. This period, though variable between glacial domains, is considerably shorter than that required to characterise the incidence of large earthquakes in other intraplate terrains. Perhaps more significantly, as is discussed later, it remains uncertain whether the nature of postglacial seismotectonics bears any relation to that of the present-day or, by inference, the future. Although this is a general issue of debate in palaeoseismology, the problem is more acute in former glaciated terrains, where the pattern, magnitude and frequency of earthquakes following deglaciation may reflect very different driving forces and a contrasting stress regime to those operating today. Thus, in Fennoscandia, present-day seismicity is concentrated around the margins of the former ice sheet, whereas on deglaciation, earthquakes predominated in the centre of the rebound dome. The association of seismicity with glacial rebound suggests that in areas experiencing diminishing rebound, the corresponding level of seismicity decreases over time. In this case, palaeoseismicity alone would offer a misleading guide to the likely incidence, magnitude and frequency of future earthquake activity. In this context, the utility of important palaeoseismological tenets, such as the earthquake loading cycle, fault-behaviour scenarios (e.g. characteristic-earthquake model) and ideas on earthquake recurrence, are questionable in glaciated terrains.

Intriguingly, perhaps the complex palaeoseismological challenges lie outside the former glaciated regions. Within and close to the limits of former ice sheets, there is an assumption that some or all of the observed postglacial deformation and seismicity has a glacial component. However, whilst rebound models show that appreciable glacial strains can accrue in the adjacent forebulge domains, glacial loading is rarely considered in those areas as a

contributory influence on the incidence of Holocene tectonism. Wu (1999) discusses the extent to which early Holocene palaeoseismic activity in southeastern USA may have a glacial component. Muir Wood (this volume) takes this further, suggesting that the great ( $M > 7$ ) earthquakes of 1811-1812 (New Madrid) and 1886 (Charleston) (Fig. 4a) may represent heightened seismic strain-release on the outward edge of a collapsing forebulge.

The difficulty in neglecting a possible glacial component is illustrated using a recent assessment of the seismotectonic significance of faults in the Upper Rhine Graben (France). The faults displace Late Cromerian deposits by 16 m and Saalian deposits by 6 m (Lemeille *et al.*, 1999). The suggestion that these displacements were coseismic slip events caused Lemeille and co-workers to argue that, according to their timescale (Late Cromerian = 400-450 ka and Saalian = 150-300 ka) the faults are characterised by an average recurrence interval of 25,000 years for  $M$  6.0-6.5 earthquakes. However, this interpretation is complicated by these ages being different from those accepted elsewhere (Funnell, 1995), and the fact that the site is within the sphere of influence of ice loads that would have developed on a large number of occasions in Fennoscandia, the Alps and the nearby Vosges massif (Seret *et al.*, 1990). Thus it is entirely possible that repeated glacial loading and unloading cycles over this period may have induced at least some of this fault movement. In this context it is necessary to question the value of time-averaged fault-slip rates and earthquake-recurrence intervals in seismic-hazard assessment. Recent studies suggest that more valuable indicators may be derived from our understanding of the local stress field operating in a region.

## 11. What does the crustal stress field tell us about the contemporary seismotectonics in glaciated domains ?

A good knowledge of the *in situ* stress field and its spatial variations is crucial to understanding contemporary geodynamic processes such as tectonic deformation or postglacial rebound, and consequently stress data are being used to calibrate and test crustal rebound models (Grollimund and Zoback, 1999). Rough estimates for the stress state of the crust comes largely from the analysis of earthquake focal mechanisms and *in situ* stress determinations, but these are often sparse or ambiguous in low-seismicity glaciated shields (Adams, 1989b; Zoback, 1992). Although broad compilations of stress-orientation data generally highlight regional tectonic stresses (Gregersen 1992), deviations from the regional stress field may reveal locally anomalous stress regimes prevail. Do such local stress anomalies reflect the activity of seismotectonic structures?

In this volume, Hicks *et al.* examine this question in the Rana region of coastal western Norway (Fig. 5), where a prominent zone of seismicity coincides with a region of elevated postglacial uplift gradients. Earthquake focal mechanisms determined from a detailed monitoring network show that the area is a pronounced stress anomaly, though the microseismicity showed no activity on a fault, the Båsmoen Fault, previously suspected of postglacial reactivation (Olesen *et al.*, 1995). Nevertheless, although other crustal factors cannot be ruled out, the results suggest that the stress anomaly, and its associated heightened seismicity, reflects deglaciation-induced crustal flexure, consistent with the modelling work of Fjelskaar (this volume). Intriguingly, other workers (Adams 1989b, Muir Wood, this volume) have speculated that the stress anomalies of the type identified in Rana may actually



represent 'stress shadows', that is, 'ghosts' of large earthquakes that occurred in prehistoric times; thereby providing a possible explanation for the modern seismic quiescence of the Båsmoen Fault. Clearly, future research needs to focus on whether stress anomalies represent an imprint of past seismotectonics or the signal of ongoing strain build up.

The paucity of *in situ* stress determinations and the potential ambiguity of focal mechanisms in low-seismicity glaciated domains has focused attention on other geological indicators of crustal strain accumulation. For example, the analysis of minor fractures offsetting modern glacially-smoothed rock surfaces exposed by historical glacial retreat allowed Norris & Cooper (1986) to compare modern postglacial strain rates to seismic strain rates along the Alpine Fault of New Zealand. Another potential indicator are 'pop-ups', postglacial structures that develop at the surface structures on artificially exhumed (e.g. quarry floors) or naturally-exposed (river valley floors) rock surfaces in response to high horizontal compressive stress (Wallach *et al.*, 1993). Wallach and co-workers argued that, in eastern North America, the mode of formation of 'pop-ups' was consistent with that of subsurface faulting inferred from earthquake focal mechanisms. Thus, their presence in an area may be indicative of the potential of that area to host moderate to large earthquakes. In this volume, Roberts describes another manifestation of high compressive strain in near-surface rocks in the form of reverse-slip offsets and axial fractures in artificial road-cut boreholes in northern Norway. Muir Wood (this volume) discusses the potential utility of such small-scale structures in elucidating the present-day strain field in former glaciated domains. However, in using such indicators, it should be kept in mind that stress fields in areas undergoing postglacial rebound are likely to be particularly complicated, with the axes of principal stress tending to switch, giving to stress regimes that change with both depth and location (Adams 1989; Talbot 1999).

## **12. Do tectonic processes or postglacial rebound dominate contemporary seismotectonics?**

A long-standing question in the study of the present-day seismicity of former glaciated regions is whether ongoing tectonic deformation or residual postglacial rebound (including both glacial loading and unloading) is the main cause of earthquake generation. In recent decades, the consensus view has been that the seismotectonics of deglaciated northern Europe and eastern North America were dominated by plate-tectonic boundary forces (Gregersen & Basham, 1989). In both Fennoscandia and Eastern Canada, the regional distribution of seismicity appears to show little correlation with the pattern of postglacial rebound, and instead seismic activity appears to concentrate along pre-weakened tectonic zones. Much of the glacio-isostatic uplift centre of eastern North America appear to be substantially aseismic, and seismic activity instead concentrates along former rift zones at its eastern continental margin (Fig. 4a) (Adams & Basham, 1989; Hasegawa & Basham, 1989). In northern Europe, most of the larger earthquakes ( $M > 4$ ) are distributed along the coastal regions, where they similarly tend to be localised in pre-weakened tectonic zones (Bungum & Fyen, 1980; Kvamme & Hansen, 1989) (Fig. 5). The interiors of these former glaciated domains, where postglacial rebound rates are greatest, are again relatively non-seismic (Slunga, 1989; Wahlstrom 1989; Gregersen *et al.*, 1991; Skordas & Kulhánek, 1992). Furthermore, the contemporary stress fields in both Fennoscandia and eastern North America do not appear to be dominated (or even perturbed) by the effects of past glaciation, and instead are most readily explained as a response to stresses exerted at the nearest plate boundary, the Mid-Atlantic Ridge (Adams, 1989b; Adams and Bell, 1991; Gregersen *et al.*, 1991). Taken together, this evidence previously had been taken to suggest a tremendous change in the

stress field in the Holocene time, from one dominated by postglacial unloading right after the Last Glacial Maximum to one dominated by present plate motion today (Mörner, 1979; Stephansson, 1989; Gregersen, 1992). But is this really the case?

The new glacial rebound models demonstrate why glaciated interiors are low-seismicity domains: rebound-induced horizontal strains are greatest at the ice margins not in the area of greatest ice thickness. For example, Mitrovica *et al.* (1994), computed crustal expansion rates of  $\sim 1$  mm/yr beneath the rebounding Fennoscandian and Laurentide shields and crustal contraction rates of  $\sim 0.5$  mm/yr in the adjacent collapsing forebulge. By comparison, horizontal crustal movements of up to 2mm/yr are predicted for the ice-marginal zones (e.g. in northeast Labrador and western Norway). The equivalent strain rates in these ice-proximal zones are likely to be significantly (perhaps two orders of magnitude) higher than the tectonic strain rates found in the continental interiors (Johnston, 1989). A spatial coincidence between steep gradients of postglacial rebound and seismicity along the northeastern margin of the former Laurentide ice sheet was noted by Hasegawa and Basham, (1989), and several papers in this volume similarly draw attention to this association in Fennoscandia (Fjelskaar, this volume; Hicks *et al.*, this volume).

Despite the amplification of horizontal crustal motions along former ice-margins, glacial rebound models predict that the largest present-day vertical motions will occur where the ice sheet had been thickest (Fig. 4b). This expectation has been confirmed by recent GPS measurements which, along with modeling results and evidence from mareograph and precise levelling, shows high rates of vertical surface displacement in the centre of the Late Pleistocene glaciation regions of Fennoscandia and North America (Scherneck *et al.*, 1998;

Argus *et al.* 1999). The seismotectonic significance of such vertical motions remain unclear. In Fennoscandia, the region of current maximum uplift rates lies close to the Lapland Fault Province (Fig. 5), though the province probably occupied an ice-margin position when the bulk of fault activity occurred. What is clear from both models and measurements of ongoing crustal motions is that rebound stresses need not be insignificant at the present day. Thus while rebound stresses in Fennoscandia are predicted to have waned progressively since early postglacial times, they remain of sufficient order to maintain crustal instability in large parts of the region (Johnston *et al.*, 1998). The occurrence of focal mechanisms of different faulting styles (normal, thrust and strike-slip) in close proximity in this region suggest local stress perturbations that may be due to residual glacial rebound (Müller *et al.* 1992). In eastern Canada, depending on the viscosity structure of the rebound model used, crustal instability may be either diminishing (low viscosity -  $10^{21}$  Pa) or increasing (high viscosity -  $10^{22}$  Pa) over Holocene times, the latter predicting enhanced earthquake activity over the next few thousand years (Wu 1998). According to Wu (1999), even with a low-viscosity rebound model it is possible to explain the spatial distribution of earthquakes in eastern Canada and Fennoscandia, their onset time, the observed stress orientations and most of the observed styles of tectonic deformation. The modelling work confirms the earlier view by Quinlan (1984) that although rebound-induced stress changes are sufficiently small (of the order of a few MPa) that they are unable to cause fracture, they are enough to trigger earthquakes on marginally stable pre-existing faults in tectonically weakened zones.

In summary, the present consensus appears to be that both tectonic forces and rebound stress are needed to explain the distribution and style of contemporary earthquakes in regions like eastern Canada and Fennoscandia (Wahlstrom, 1993; Arvidsson & Kulhanek, 1994; Wu, 1999; Wu *et al.*, 1999). However, other forces may also contribute. In Fennoscandia, recent

studies have argued that although plate-boundary tectonic stresses drive the regional compressive stress field, these far-field tectonic stresses are themselves insufficient to account for the current level of seismicity, and, therefore, other regional or local mechanisms must operate, such as flexural stresses from sedimentary loading and/or glacial rebound (Byrkjeland *et al.*, 2000; Fejerskov & Lindholm, 2000). In this volume, Fjeldskaar argues that the bulk of observed seismicity in Fennoscandia can be explained by a model of postglacial rebound that is consistent with the observations of deglaciation, palaeoshoreline tilts and the present rate of uplift. However, the coincidence of pronounced seismic activity in areas with anomalously high uplift rates suggests that glacio-isostatic movements are locally overprinted by a weak tectonic component. Fjeldskaar notes that not all of this component need be tectonic, since the areas of anomalous uplift occur in the topographically elevated zone of western Fennoscandia where enhanced glacial erosion during the Late Pliocene and Early Pleistocene may similarly cause additional compensatory uplift. The effect of topography is neglected in the most current Earth model, as is lateral variations in lithospheric thickness and/or asthenospheric viscosity, but both factors are likely to be important in simulating postglacial-rebound responses at the mountainous rifted margins of deglaciated shields, such as that of western Norway and of Baffin Island (Arctic Canada) (Hasegawa and Basham, 1989; Muir Wood this volume).

In this volume, Muir Wood argues that current Earth models of postglacial rebound neglect an additional factor: tectonic strain-energy accumulating during the period that the ice-sheet was in existence. The interaction of the postglacial rebound strain field with the prevailing tectonic strain field will cause areas of constructive interference (where crustal instability and seismicity is promoted) and areas of destructive interference (areas of aseismicity). Although seismicity generally occurs around the margins of the former ice sheet, Muir Wood predicts

that the highest magnitude earthquakes will occur in those quadrants of the former forebulge where the maximum horizontal compressive stress coincides with the direction of tectonic strain polarisation. Adams (1989a) had proposed such a similar conceptual model for eastern North America, but Muir Wood argues that the model is also capable of explaining the bulk of the seismicity observed across northern Europe (Britain and Fennoscandia).

### **13. The modulation by glaciers of the mountain building process**

The question of whether tectonic uplift or adjustment to glacial erosion might account for zones of anomalous present day uplift in Fennoscandia (Fjelskaar, this volume, pp - ) touches upon an issue of major current debate - how climate, tectonism, and topography interact in the mountain building process (e.g., Molnar and England, 1990, Small, 1999; Whipple *et al.*, 1999). Part of this debate centres around the role of glaciers in setting the form of orogenic topography and in their efficiency as erosional agents. For example, studies in the Southern Alps of New Zealand (Kirkbride & Mathews, 1997) suggest that ca. 70 ka is needed for glaciers to erode recognisable parabolic profiles along V-shaped river valleys, while it will take ca. 320 ka to erode mature glacial troughs.

Hitherto, research on glaciers as a driving force for mountain building has focused on how fluvial erosional systems in orogenic terrains are influenced by climate, tectonism, and geomorphology (Koons, 1989; Whipple and Tucker, 1999, Willett, 1999). This tendency has been prompted, in part, by the fact that river erosion tends to predominate in active orogenic belts at present, though the role of other processes is now receiving some consideration. However, modern analogue studies must be of limited value as the forcing factors operating

during the colder periods of the Quaternary were different from those of today, and many active orogenic belts have been extensively glaciated for significant periods of time. The role of glacial erosion on the mountain-building process therefore needs to be considered more fully (e.g., Porter, 1989; Brozovic *et al.*, 1997).

One orogen in which erosion by glaciers played, and continues to play, an important role is in southern Alaska (Meigs and Sauber, this volume). In this setting, mean orogen-scale erosion involves primarily glacial erosion of bedrock as well as erosion in areas of the landscape that are ice-marginal and are deglaciated at glacial minima. Former erosion is likely to be most efficient near the equilibrium line altitude (ELA) of the glaciers where the ice sliding velocity and the glacier thickness is greatest. Oscillations between glacial and interglacial climate control ice distribution, thickness and gradient, which in turn, modulates the locus, mode and rate of erosion in the landscape. To quantify the form of the landscape in this glaciated orogen, Meigs and Sauber characterized topographic form, hypsometry, mean elevation, and slope regionally using a digital elevation model of southern Alaska. The range scale influence of glacial erosion was estimated by comparing landscape characteristics with the location of the current and LGM ELA's where ice-bedrock interface erosion is inferred to be highest. These data, along with recent estimates of ongoing deformation patterns from GPS geodetic measurements, have served as a basis to explore the linkages between the modern plate-boundary deformation field, long-term patterns of crustal deformation, and the temporal and spatial distribution and fluctuation of glaciers.

## Conclusions

Emerging studies show how glacial dynamics may modulate the deep-seated viscoelastic response of the Earth, the nature and incidence of upper-crustal faulting and earthquake generation and even the mountain-building process itself. Such bold contentions arise from a new generation of postglacial rebound models that show how the viscoelastic response of the mantle can exert significant horizontal and vertical stresses both within and outside the limits of former ice sheets. These glacial rebound models, however, use estimates of Earth rheology that could be better constrained with additional spatial and temporal data; and high-resolution palaeoshoreline and sea-level change data will continue to be critical in serving this purpose (e.g., Gehrels *et al.*, 1996; Tackman *et al.*, 1998). Additionally, such data may provide information on local crustal movements (Dyke, 1993; Gehrels, 1994; England, 1997).

In the absence of absolute age control, high-resolution morpho-stratigraphic deglaciation chronologies are critical for constraining the timing of fault movement in many former glaciated domains. However, this is not restricted to the tectonically quiescent glaciated shields, since in tectonically active settings too, glacial landforms or stratigraphies may provide the basis of fault-movement histories (Sutherland & Norris, 1995; Brown *et al.*, 1998). This, together with the difficulties in discriminating tectonic faulting and palaeoseismic deformation from glaciogenic or cryogenic phenomena, will require a far closer collaboration between seismologists and earthquake geologists and Quaternary scientists and glaciologists.



Improved understanding of past seismic activity has rewards not only for seismic-hazard and radioactive waste-disposal research, since earthquakes are increasingly regarded as a dominant control on sedimentation patterns along former glaciated continental margins (Laberg & Vorren, 1995; Vorren *et al.*, 1998; Hunt & Malin, 1998). These sedimentation patterns, in turn, provide one means of assessing the long-term contribution of glacial erosion to marginal uplift (Vagnes *et al.*, 1992). Even away from the centres of glacial loading, the interplay between tectonic movements and erosion-driven isostasy has been interpreted as important controls on valley incision and river-terrace formation (Maddy, 1997). Such interplay is likely to be even more acute along tectonically active glaciated uplands. In short, the interactions between ice sheets, crustal deformation and seismicity operate at a range of scales and in a dynamic and complex fashion, and collaboration between Quaternary scientists and geophysicists and tectonicians is essential in unravelling them.

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## Table Captions

Table 1: Summary of the attributes of documented postglacial faults within the Lapland Fault Province (from Dehls *et al.* 2000). The major faults are NE-SW trending reverse faults and occur within a 400 x 400 km large area in northern Fennoscandia. The scarp height/fault length ratio is generally less than 0.001. Moment magnitudes are calculated from the fault offset and length parameters utilising relations of Wells and Coppersmith (1994). Note that other studies have suggested that moment magnitudes on the Pärve and Lansjärv faults may be as high as 8.2 and 7.8, respectively (Arvidsson 1996, Johnston, 1996).

## Figure Captions

Fig. 1: Conceptual model for near-surface postglacial stresses during deglaciation. The exploded block model illustrates how the superimposition of changing glacial flexural stresses (light grey arrows) on a uniform regional tectonic stress field (black arrows) results in contrasting resultant stress states (dark grey) and different styles and orientations of stress-relief phenomena at the ice margin (A and D), the forebulge (B and E) and the undeformed foreland (C and F). Modified from Adams (1989a) after Walcott (1970). Note the considerably exaggerated vertical scale difference between the lithosphere thickness and the ice sheet thickness.

Fig. 2: A simplified loading history for Fennoscandia during the last (Weischelian) major glaciation (from Talbot 1999, fig. 6a). Ice thickness appears above the time axis and consequent crustal depression appears below. The model illustrates how rapid vertical



ice loading and unloading changes lead to far slower crustal depression or rebound, and that the scale of crustal response is considerably less than the imposed ice-mass fluctuations, reflecting the deeper mantle flow.

Fig. 3: A diagrammatic representation of the impact of glacial loading (a) and unloading (b) on the crust in a region with a compressive stress regime. Modified from Muir Wood (1989) and Fenton (1992). Note the considerably exaggerated vertical scale difference between the crustal thickness (~100 km) and the ice-sheet thickness (~3 km).

Fig. 4: (a) Distribution of seismicity (open and solid circles) in eastern North America in relation to the main ice limits (solid line) and principal ice divides (dashed lines) at the Last Glacial Maximum. The dates and magnitudes of large or notable earthquakes are given. The comparative aseismicity of the Laurentide glacial interior contrasts markedly with the apparent seismic expression of its former margins. Despite the coincidence of earthquakes with the ice margins, greatest stress directions inferred from focal-mechanism P axes of recent earthquakes appear to be aligned parallel to the regional direction of maximum horizontal compression ( $\sigma_{Hmax}$  - thin grey arrow pairs) imposed by tectonic stress (Mid-Atlantic ridge push). Black arrows show horizontal motions (mm/yr) predicted by the postglacial rebound model of Peltier (1996) for sites at which there is new geodetic (very long baseline interferometry and satellite laser ranging) control (Argus *et al.*, 1999). The arrows illustrate the outward radial pattern of crustal motion around the former ice sheet, though GPS results indicate that the rates are probably too high; sites beneath the former margins of the Laurentide ice sheet are currently moving laterally away from the ice-sheet centres at <1.5 mm/yr (Argus *et al.* 1999). Inferred directions of  $\sigma_{Hmax}$  at early postglacial

time (deduced from small thrust faults) are shown in the short grey inward-pointing arrowheads (from Adams 1989), and appear to be parallel to the radial glacial stresses. The open squares show the locations discussed in the text where late Quaternary faulting is observed or proposed: the Apsy Fault (A); the Ungava Peninsula (U); and Peel Sound (P). Seismicity is after Johnston (1989b), augmented by new data from Canada from Adams (1996). Ice limits are after Dyke & Prest (1987), Andrews (1991, fig. 1) and England (1997).

(b) NW-SE cross section across North America comparing model predictions of vertical crustal motions with observed motions from the GPS results of Argus *et al.* (1999). The results support the expectation that highest rates of vertical surface displacement occur in the former centre of the Late Pleistocene ice sheet.

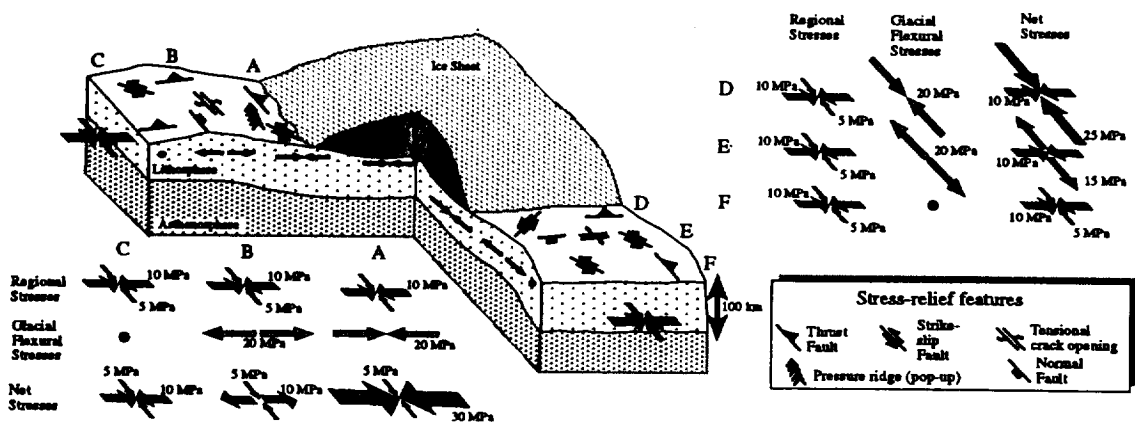
Fig. 5. Location of earthquake epicentres and postglacial faults in the Lapland Fault Province of northern Fennoscandia. Numbers indicate faults whose parameters are given in Table 1. The location of the Båsmoen Fault, a suspected postglacial fault at Mo i Rana (Hicks *et al.*, this volume) is also shown. Elliptical zone defined by dashed line shows the region in which GPS measurements (BIFROST project) indicate vertical surface uplift rates  $> 10$  mm/yr (Scherneck *et al.*, 1998).

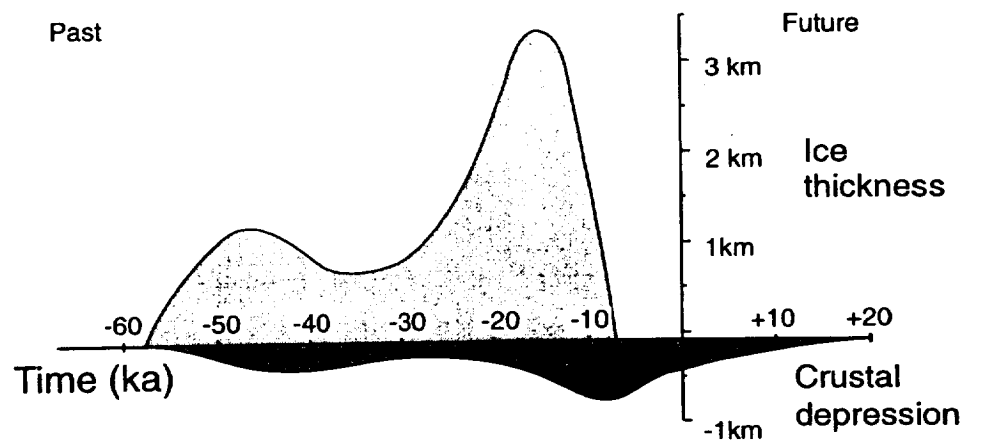
Fig. 6. Changing styles of faulting in Fennoscandia with changes in the pattern of glacial loading and unloading during deglaciation. Normal faulting is predicted in the white area, strike-slip faulting in the light-grey region and thrust faulting in the dark-grey region. From Johnston *et al.* (1998, fig. 13).

Table 1: Summary of the attributes of documented postglacial faults within the Lapland Fault Province (from Dehls *et al.* 2000). The major faults are NE-SW trending reverse faults and occur within a 400 x 400 km large area in northern Fennoscandia. The scarp height/fault length ratio is generally less than 0.001. Moment magnitudes are calculated from the fault offset and length parameters utilising relations of Wells and Coppersmith (1994). Note that other studies have suggested that moment magnitudes on the Pärve and Lansjärv faults may be as high as 8.2 and 7.8, respectively (Arvidsson 1996, Johnston, 1996).

No.	Fault	Country	Length (km)	Max. Scarp Height (m)	Scarp height / fault length ratio	Trend	Type	Moment magnitude	Reference
1.	Suasselkä	Finland	48	5	0.0001	NE-SW	reverse	7.0	Kujansuu, 1964
2.	Pasmajärvi- Venäjärvi	Finland	15	12	0.0008	NE-SW	reverse	6.5	Kujansuu, 1964
3.	Vaalajärvi	Finland	6	2	0.0003	NW-SE	??	6.0	Kujansuu, 1964
4.	Pärve	Sweden	150	13	0.0001	NE-SW	reverse	7.6	Lundquist & Lagerbäck, 1976
5.	Lainio- Suijavaara	Sweden	55	30	0.0005	NE-SW	reverse	7.1	Lagerbäck, 1979
6.	Merasjärvi	Sweden	9	18	0.002	NE-SW	reverse	6.3	Lagerbäck, 1979
7.	Pirtimys	Sweden	18	2	0.0001	NE-SW	reverse	6.5	Lagerbäck, 1979
8.	Lansjärv	Sweden	50	22	0.0004	NE-SW	reverse	7.1	Lagerbäck, 1979
9.	Burträsk- Bastuträsk	Sweden	60	~10	0.0002	NE-SW	??	7.1	Lagerbäck, 1979
10.	Stuoragurra	Norway	80	7	0.0001	NE-SW	reverse	7.3	Olesen, 1988
11.	Nordmannvik	Norway	3	1	0.0003	NW-SE	normal	5.7	Tolgensbakk & Sollid, 1988

Stewart, Sauber and Rose - Fig. 1 (Adams - Freehand 5.0)





Stewart, Sauber and Rose - Fig. 3 (Muirwood2 - Freehand 5.0)

